DEEP ICE CORE SITE EVALUATION FOR RIDGE B/C, WEST ANTARCTICA

by

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ABSTRACT

An understanding of the nature and extent of changes of the West Antarctic ice sheet are important in establishing its stability to past and future changes in climate and sea level. Two competing views of its past thickness and present overall stability exist. One view, championed Denton and Hughes, maintains that the ice surface was up 1600 m thicker at - 18,000 years and that the ice sheet is presently undergoing a complete collapse. Drewry and Robin disagree, instead they suggest that a relatively stable ice sheet that has been modified by a moderate surface lowering of at most 600 m is a more appropriate view.

The approach used in this study was to apply a physically based kinematic flow model of ice to the two competing theories and predict likely core stratigraphy. This thesis tests whether a program of ice coring atop an inter-stream ridge (specifically ridge B/C) could discriminate between such ice sheet models.

The results of modeling past ice flow to the small dome on ridge B/C (Siple Coast, West Antarctica) indicates that a properly designed ice core program could resolve changes in ice thickness since -18,000 years. Such a core could, in addition, discern changes in the ice divide position related to the emplacement of the ice dome.

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DEDICATION

To my father William O. Graham 1925-1988 and my mother Carol T. Graham for their love and their endurance. Thank you.

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INTRODUCTION

Stored within the ice sheets of Antarctica and Greenland is perhaps the most detailed record of climatic change available for study. During the last three decades five major ice cores have been extracted at sites Dye 3 and Camp Century in Greenland, and at Byrd Station, Dome C, and Vostok in Antarctica. These deep cores have opened this record to investigation and their extensive analyses have contributed greatly to our detailed climatic knowledge of Late Wisconsin and Holocene time.

Recently, a great deal of interest has been aroused over coring at Vostok. Here, with drilling completed only partially through the ice sheet, a core containing the first continuous and complete record of one glacial cycle (approximately 160,000 years) has been removed (Jouzel et al 1987). Soon, planned Danish and American cores in north-central Greenland may produce records of similar length and thus permit correlation of the climatic responses of northern and southern hemispheres over this time scale.

The actual record stored in an ice core is the result of a rather complex jumble of factors related to climatic and ice sheet changes, and the normal flow of ice. The task of sorting them all out often proves to be daunting. To begin, each successive layer of snow that accumulates on the surface of a glacier brings with it a record of the climatic conditions that existed at the time and place of deposition. This record may take many forms, for example total gas, carbon dioxide, and micro-particle dust; however, isotopic profiles are generally regarded to be the most accurate, detailed record of climatic change presently available (Robin 1983).

Specifically, these isotopic profiles refer to the ratios of the stable isotopes of hydrogen and oxygen (deuterium and O^{18}) to normal hydrogen and oxygen reported as deviations from a standard arbitrary mean called SMOW (for Standard Mean Ocean Water) plotted as function of depth. While, these two records exist independent of one and other, they tell essentially the same story and may be used almost interchangeably. Both are indicators of air temperature at the time of deposition. But, since air temperature is linked to elevation by the adiabatic lapse rate (for dry air $\Gamma_d \approx 10^{\circ}$ C km^{-1} from Houghton 1977) these values have a strong elevational component as well. Hence, to deduce the magnitude of a given temperature (climate) change implied in a core profile one must know the elevation of the ice particle at deposition.

The record of elevation of deposition of an ice particle is stored in the total gas composition of glacier ice. Total gas data has been shown to be particularly sensitive to air pressure at pore close off time (Robin 1983). Air pressure is itself linked to sea level changes and so an interpretation of total gas data must account for this effect. Together, under ideal conditions, the total gas and isotope profiles can indicate climatic change and the elevational response of an ice sheet to this change.

In order to understand the relationships between cause and effect; climatologists need to document not only the changes themselves but when they occurred as well. To do so, requires that the relationship between core depth and age be established. If conditions are favorable for their production, annual isotopic layers can provide this relationship. However, ice flow thins these seasonal layers and isotopic diffusion tends to otherwise obliterate them with depth. In addition, those layers lost in the coring process or years not recorded due to unusually low annual accumulation prove difficult to assess (Johnsen 1972). So while in principle, one could simply count them like rings on a tree; in practice, they tend to be useful as constraints in modeling.

Numerical flow modeling of ice movement can overcome some of these shortcomings. The measured isotope, and total gas profiles, and layer thinning rates can be used as constraints in particle trajectory models that calculate the effects of down slope movement, surface elevational changes and layer thinning. Implicit within this type of calculation is an age-depth relationship.

In general, numerical modeling has proven particulary successful when applied to cores from ice sheets of Greenland and East Antarctica. These continental ice masses are thought to be generally stable in regard to key features such as flow lines, ice domes and divides. Their stability affords the modeler a greater margin of confidence when assessing other possible types of changes in ice sheet configuration and ice flow.

In many ways, the Camp Century core represents glaciological nirvana having good isotopic data (including annual layers) and general ice sheet stability. The ability for such a core to resolve relatively minor changes is high. For example, modeling conducted by Bolzan and Whillans on the Camp Century ice core in western Greenland was able to date a change in flow regime. This change amounts to a collapse of a portion of the ice sheet in the Melville Bay region and the formation of an east-west trending ridge north of Camp Century (Bolzan and Whillans in prep).

While annual layer data are important they are not absolutely essential. For example, despite a lack of them in the Vostok core, an age verses depth relationship has been established by flow modeling that is generally accepted. Here again the stability of the ice sheet is crucial as ice flow in this region is generally regarded not to have changed greatly.

Attempts to model the flow to the Byrd ice core in West Antarctica have met with considerably less success. Two decades after the core was drilled, the age-depth relationship is still only tentative known (Budd and Young 1983, Robin 1983, and Whillans 1983, and Whillans 1979). Some of the problem stems from the dearth of annual layer data in the measured $\delta^{18}O$ profile. Byrd, in marked contrast with Camp Century, shows almost no reliably seasonal variation (Johnsen et al 1972).

However, couple this problem with concerns about West Antarctica's overall stability (Hughes 1973, Weertman 1974, Thomas 1976, Denton and Hughes 1981), which can and do imply large changes in not only surface elevation but also in ice divide position, and the result is a certain amount of skepticism for any proposed depth versus age relationship that is based solely on flow modeling.

Despite this, the information contained within the Byrd core remains of particular value because in it ice deposited during Wisconsin-Holocene transition occurs between 800 m and 1100 m above bedrock. At these depths irregularities in core data due the bed should be minimal (Robin 1983). However, the limiting factor in the quality of this information is our ability to place this into it's proper context, which is itself subject to our knowledge of changes in flow conditions. The dilemma is as follows.

The West Antarctic ice sheet is by definition a marine ice sheet, it's bed, even when the effects of isostasy have been accounted for, lies vastly below present mean sea level. While, the predominate drainage pattern of this region is through fast flowing ice streams (see figures 1 and 2) that debouch either directly into the ocean, or into floating ice shelves (Hughes 1977, Muszynski and Birchfield 1987). Like continental ice, marine ice is subject to interactions with the atmosphere; however, marine ice is additionally subject to sea water conditions. Variations in sea level, temperature, and circulation patterns all have potentially large impacts on marine ice systems (Muszynski and Birtchfield 1987).

Thus, most workers agree that due to the presence of these additional feedback systems, to which continental ice is less subject, the response time of marine ice to climatic variation is significantly quicker than that of continental ice. Some, however, carry this point one step further and suggest that marine ice is inherently unstable and therefore unable to exist in the steady-state (Hughes 1981).

As editors of the Last of the Great ice sheets, Denton and Hughes propose such a view (Denton and Hughes 1981). Arguing from primarily a glacial geologic perspective they suggest that the West Antarctic ice Sheet was greatly thicker (in places more than 1600 m) during the last glacial maximum than it is now. From striae on exposed bedrock, and marginal glacial deposits they conclude that the ice sheet was then fully grounded and extended virtually to the continental edge. A well recognized 120 m eustatic sea level drop is cited as the cause of this grounding.

Others, Robin and Drewry (1983), Robin (1985) for instance, contend that the Holocene thinning was more moderate and relegated primarily to the ice sheet's margins. Based on numerical modeling they assert that the 120 m sea level drop was of insufficient duration to fully ground the West Antarctic ice sheet. Instead they suggest more minor advancement of the grounding line accompanied by a thickening of at most 600 m.

Earlier calculations done by Weertman that suggest West Antarctica is stable tend to support this view, however, he concedes that this balance is, at best, a tenuous one (1974).

Modeling conducted on the Byrd core by Whillans (1977, 1979, 1983) shows the Byrd Station Strain Network (BSSN) region to have been at or near steady-state for the last 30,000 a but that a 200 m surface lowering is consistent with internal radar reflections. These conclusions generally fit the predictions of Drewry and Robin. But, Denton and Hughes (1981) counter that these results are not totally inconsistent with their view either. They suggest that their model ice sheet may have collapsed before achieving it's equilibrium thickness in the central region; and so the data may be construed to fit their model as well.

Thus the argument over West Antarctica breaks into basically two schools of thought centered on the issues of the ice sheet's long term stability and past configuration (see figure 3). The focus of this paper is to examine the potential for a deep ice core drilled on Ridge B/C to resolve past changes in ice sheet configuration.

Ridge B/C is a likely candidate for such a study for several reasons. In general, the Siple Coast is a region that both models of the past ice sheet predict the greatest magnitude of change in ice surface elevation during the Holocene transition and these change should manifest themselves in any core drilled in this area. Secondly, the regional ice divide is well constrained to the nunataks of the Whitmore mountains and thus the flowline is as well. Thirdly, ridge B/C lies between two ice streams that are very different (ice stream C is inactive while B shows remarkable velocities) and a core in this region will serve to unify already existing data sets.

The approach used in this study was to apply a physically based kinematic flow model of ice to the two competing theories and predict likely core stratigraphy. This comparison allows an assessment of the potential of a core drilled on ridge B/C to resolve past flow configurations.

THE NUMERICAL MODEL

The following is a condensed description of the particle trajectory model; those wishing a more complete version please refer to Bolzan and Whillans (in preparation).

The particle trajectory model employed in this study is based on the equation of continuity. Which is not an equation of motion but merely a relationship that must hold for the conservation of mass (Marsden and Tromba 1981). It is employed to provide a mean-through-thickness velocity at any time and any position along the flowline. From this, the horizontal and vertical velocity components are then computed using a parameterization of the shape of the velocity profile and the assumption that flow is planar. This information is sufficient to describe particle trajectories both forward and backward in time. In this study the calculation starts at selected core depths then proceeds backward along the trajectory up the flowline until it intersects with the surface of the ice sheet at the elevation and position of origin.

Continuity can be expressed:

$$\frac{\partial}{\partial x}\left(H\bar{u}_x\right) + \frac{\partial}{\partial y}\left(H\bar{u}_y\right) = \dot{b} - \frac{\partial H}{\partial t} \qquad (1)$$

where H is the ice sheet thickness, $\bar{u}_x$ and $\bar{u}_y$ are the mean-through-thickness velocity components in the horizontal directions. The accumulation rate in ice equivalent thickness is $\dot{b}$, and, time, t, is positive towards the future.

Integration of this equation is much simplified by defining a coordinate system so that the x-axis is parallel to the flowline of interest. The z-axis is positive upward from present mean sea level. The y-axis is then mutually perpendicular to both these axes. This choice of coordinates allows the use of a simplifying substitution for the y-component of velocity u_y:

$$\bar{u}_y = y\frac{\bar{u}_x}{R} \tag{2}$$

where, R, is the distance from the present x value to where neighboring velocity vectors intersect. More importantly, since flow is generally perpendicular to elevation contours (Whillans, Bolzan and Shabtaie 1987) R is also the radius of curvature of the contours. Refer to figure 4 for a geometric proof of this relationship. Making this substitution for $\bar{u}_y$ (equation 2) into the differential equation (1) for $\bar{u}_x(x,y=0,t)$ we get:

$$\left(\frac{\partial H}{\partial x} + \frac{H}{R}\right)\bar{u}_x + \frac{\partial}{\partial x}H\bar{u}_x = \dot{b} - \frac{\partial H}{\partial t} \tag{3}$$

This differential equation is then solved under the boundary condition that the horizontal velocity is zero at the ice divide. When solved (done in this model using finite differencing) it provides the mean through thickness velocity at any point along the flowline at any time for a given ice sheet configuration, accumulation rate and thickening rate. The parameters H(x,t), R(x,t), $\dot{b}$(x,t) are all that are required for this calculation.

In order to compute particle trajectories, the depth variation of the velocity components must be specified. In this model a simplified parameterization for the depth variation of velocity is employed. It is based upon the constitutive relationship and accounts for effects due to depth increasing shear stress, depth variation in temperature and crystal anisotropy. This is the same approach used by Whillans (1979) for flow leading to Byrd core Antarctica and similar to that used by Dansgaard and Johnsen (1969) for flow to Camp Century. The shape function in this model, ψ, is defined by:

$$u_i(x,y,z) = \bar{u}(x,y)\,\psi(x,y,z) \qquad i = x,y \qquad (4)$$

Implicit in this approach is the assumption that flow is planar, or in the same direction at all depths.

While similiar to Dansgaard and Johnsen's approach, in which the shape function divided the glacier into an upper and lower regions with the value in the upper region constant and decreasing linearly from this value to zero at the bed in the lower region, this model uses a more flexible two parameter function.

$$\psi(x,y,z) = n\left[1 - \xi\left(\frac{z_s - z}{H}\right)^{p+1}\right] \qquad n = \frac{p+2}{p+1+\xi} \qquad (5)$$

here $z_s(x,y)$ is the position of the ice sheet surface, z is the depth, represents the proportion of flow due to internal shear and is a factor to ensure = 1 as required by equation (4). In its most general case may vary from 0 to 1 in both time and distance; however in this study is held to be 1 corresponding to no bottom sliding. The variable p may also take a range of values from 0 to any positive real number; however, a value between 1 and 3 typically provides a good fit to measured profiles. Again the value may vary both in time and distance; however, for our purposes it was held at p = 2. Figure 5 shows as a function of depth for various values of p.

The horizontal velocity components may then be described by the composition of $\bar{u}_x$ and $\bar{u}_y$ to the shape function, ,as follows:

$$u_x(x,z,t) = \bar{u}_x(x,t) \quad \psi(x,H(x,t)) \tag{6}$$
$$u_y(x,z,t) = y/R \ \bar{u}_x(x,t) \ \psi(x,H(x,t))$$

Note that, appropriate to the definition of the coordinate system u_y is zero along the flowline where y = 0.

The vertical velocity is computed by integrating the vertical strain rate. Since ice is incompressible, the vertical strain rate is linked to the horizontal strain rates by (Paterson 1969 p.70):

$$\dot{\epsilon}_{zz} = \frac{\partial \bar{u}_x}{\partial x} - \frac{\partial u_y}{\partial y} \tag{7}$$

which is essentially a restatement of the continuity relationship. Substituting from equation6 we get:

$$\dot{\epsilon}_{zz} = -\psi\left(\frac{\partial \bar{u}_x}{\partial x} + \frac{\bar{u}_x}{R}\right) - \frac{\partial \bar{u}_x}{\partial x}\psi \tag{8}$$

Where $\bar{u}_x$ is the solution to equation 3 when R, $\dot{b}$, H are all specified as functions of position and time.

Integrating equation 8 we get:

$$\begin{aligned} u_z(x,z,t) = {} & u_{zb}(x,t) + \left(\frac{\partial H}{\partial t} - \dot{b} + \bar{u}_x\frac{\partial H}{\partial x}\right)\left[1 - n\zeta\left\{1-\xi\zeta^{m}(m+1)^{-1}\right\}\right] \\ & + \bar{u}_x n\left[(1-\xi)\frac{\partial z_b}{\partial x} + \xi(1-\zeta^{m})\frac{\partial z_s}{\partial x} + \xi m\frac{\partial H}{\partial x}\left\{1-\zeta^{(m+1)}\right\}(m+1)^{-1}\right] \\ & + \bar{u}_x H\frac{\partial \xi}{\partial x}(1-\zeta^{m})n^2(m-1)^{-1} + \bar{u}_x H\xi\zeta\frac{\partial p}{\partial x}\left\{\zeta^{m}\left(1-[m+1-\xi]\ln\xi\right)-1\right\}\frac{n^2}{\{m+1\}^2} \end{aligned}$$

Where $\zeta = \left(\frac{z_s - z}{H}\right)$ is the reduced depth and $m = p + 1$ to ease computation. Basal melting and effects due to isostasy and erosion have been neglected so that $u_{zb}(x,t) = 0$. Since ξ and p are assumed to be constant in both time and position the last two terms are also zero.

This model is designed to predict core stratigraphy due to ice flow and changes in ice flow. It is used to calculate four core quantities, layer thickness, the elevation of origin of an ice particle, the layer age, and the initial x-position of the particle of ice. In application of this model to existing core data the first two quantities are directly measurable and would then be used as the constraints; while, age and initial position are predictions. However, in the absence of actual core data, the application of the model to this study was somewhat different.

The approach was to model the two major views on the evolution of the West Antarctic ice sheet and ascertain whether data from a core drilled on Ridge B/C could, in theory, resolve the differences. The constraints in this approach were thus limited to the past geometry predicted by the individual model and to a lesser degree the ice sheet's evolution to the present configuration.

STEADY STATE

The steady-state scenario represents a baseline model of the Siple Coast. In it, as it's name implies the ice sheet was assumed to always maintained it's present configuration (see figure 2). Implicit in this assumption is the permanence of such large-scale features as the ice streams and ridges. Modeling of flow under assumed steady-state conditions provides a backdrop on which to view the other more dynamic scenarios.

A hypothetical bore hole was chosen to lie on the south side of ridge B/C at approximately 610 meters elevation. A 13 km long flowline was mapped back to the local ice divide; the summit of ridge B/C (elev. 650 m). While, ideally an ice core would be drilled atop the ice divide where the horizontal velocity is zero; this model encounters difficulty when the initial x-position of an ice particle nears the ice divide. One of the boundary conditions for this model is that at the ice divide the mean-through-thickness velocity is zero. This quantity appears in the denominator of one of the terms and division by zero produces an error. The 13 km displacement is an attempt to place the bore hole as close as possible to the present local ice divide on Ridge B/C without encountering this problem in the more dynamic scenarios. Unfortunately, the steady state model still encounters this problem limiting the deepest calculated layer to 800 m.

The relevant geometry was specified; the spreading parameter, surface and bed slopes were all based on values obtained from Shabtaie, Whillans and Bentley (1987). The accumulation rate of 150 mma^{-1} of ice equivalent is the value that Whillans, Bolzan, and Shabtaie (1987) report as a regional average. Following the conclusion of Rose; the ice was modeled frozen the bed (1979). No input parameters were time dependent.

The results obtained from this configuration are reasonable. The calculated surface velocity of 1.7 ma^{-1} is in line with that measured by Whillans, Bolzan, and Shabtaie of 3 ma^{-1} considering that measured value has a flow line of about 35 km (1987). With the model set up to for such a flow line, the calculated surface velocity becomes about 4 ma^{-1}. This agreement of the measured and calculated velocity coupled with the fact that its direction is generally towards central West Antarctica suggests that the ridge is a local ice divide capable of sustaining flow.

The core predicted properties are limited to the top 800 m of the 1000 m ice thickness because of the aforementioned problem. At that depth the ice has a calculated age of -18,000 a and an initial elevation of 642 m (figures 6 and 7). Although, the plots of initial position and initial elevation versus depth appear linear they are, strictly speaking, not.

THE DYNAMIC SCENARIOS

As different as the Drewry and Robin scenario is from that of Denton and Hughes they are remarkably alike as far as model input parameters for this flow line are concerned. Since both scenarios evolve to the same end point they share at least those input parameters that govern the present ice sheet. Both predict that a thicker ice sheet in this region implies a more non-divergent flow pattern then is now observed.

The flow line for times older than -18,000 a begins at the Whitmore Mountains. While, the flow line for times younger than -4,000 a is the same 13 km flow line in the steady-state scenario. The divide is assumed to change from the Whitmore Mountains (x = -550 km) to ridge B/C (x = -13 km) between -18,000 a and -4000 a. The mechanics of this movement are probably more involved than this, and judging from the convoluted contours on the saddle of ridge B/C is likely to be related to the recent capture of flow lines by ice stream B and C. The effect of this is that flow from the Whitmore mountains would have been sustained longer and flow from the ridge would be a more recent feature than is modeled. However, since no core evidence to the contrary exists, the simplest explanation is for the change to have occurred in a linear fashion timed with the other ice sheet changes.

A similar argument for the mechanics of the change in the spreading parameter exists. Again linear changes are assumed for simplicities sake. Past non-divergent flow throughout region as implied by the two scenarios is supplanted by a flow that is very divergent at the new divide but still relatively non-divergent at the bore hole as implied by present topography (figure 2).

Since the same flow line is predicted by both models and the effects of isostasy have been ignored the same bed is also employed (see figure 8). Isostasy has been ignored because crustal rebound rates are at best difficult to access because their calculation ultimately depends on knowledge of past ice thickness; and hence, becomes a bit of a circular argument. Besides, they are of secondary importance because the driving stress of a glacier is related primarily to its surface slope which is specified elsewhere in the model.

The glacier was assumed frozen to its bed to conservatively allow for the greatest internal shear. Likely the basal boundary conditions are much more complex then this.

The specified bed reflects the major features seen in radar maps produced independently by Rose (1979) and Shabtaie, Whillans and Bently (in press). These are: a topographic high under the summit of ridge B/C at about -450 m that trends downward into a -1500 m deep trench approximately at the base of the Whitmore mountains. From here the slope abruptly changes in both magnitude and sign and rises to about 2200 m at the summit of the Mt. Chapman. Note, the first 13 km of the bed are representative of the final flow line off ridge B/C. This has the effect of underestimating the ice thickness in this area by about 50 m, while maintaining the proper bed relationship for those times greater then -4000 a.

The accumulation rate was held both regionally and temporally constant at a value of 150 mm^{-1} per year (Whillans, Bolzan and Shabtaie 1987). Likely, it was much different during the Wisconsin, but since we are most interested in changes occurring after about -18,000 a this should not provide a major point of conflict.

Both scenarios predict that the greatest thickness occurred some time near -18,000 a and that the ice sheet collapsed soon after that (Denton and Hughes 1981, Robin and Drewry 1983). Neither are explicit on the emplacement of the present flow configuration. The Denton and Hughes collapse implies that ice sheet is still in transition; while Drewry and Robin suggest that the ice sheet is stable. For the sake of the model we will say that ridge B/C has been a local ice divide since -4000 a roughly corresponding to the end of the sea level rise (see Denton, Hughes 1981). This is a tenuous assumption and shall be dealt with later.

Where they differ is, of course, in the area of surface elevation changes and the past slopes implied in these changes. Drewry and Robin suggest that the ice sheet has been affected by relatively modest changes in surface elevation. Their model predicts a maximum thickening of about 600 m in the region of the ice streams which translates to roughly 400 m atop the present ice ridges, while in the Whitmore mountains they suggest a value of about 200 m. Adding these values to the present elevations (see figure 2), we calculate a past surface slope of about 2.46 x 10^{-1} based on a 550 km long flow line.

Denton and Hughes on the other hand suggest a more radical change in West Antarctica. Their model predicts a past surface about 1600 m above the present summit of ridge B/C and about 600 m more near the Whitmore Mountains. Based on the values we calculate a slope of about 1.10 x 10^{-1} on the same 550 km flow line (figure 8).

NEW MODEL

Comparing the implied resultant core data produced by these two scenarios we get depth versus elevation profiles that are clearly distinguished from one and other (figure 10). The depth versus age relationships also diverge (figure 9); more importantly though, the depth of the -18,000 a layer is 675 m for Denton and Hughes, and 725 m for Robin and Drewry of a 1000 m thick core. This means that all recorded changes in ice sheet configuration occur relatively high in the core and therefore have a more likely chance of remaining interpretable.

As mentioned earlier there is a problem with this collapse model because it links the movement of the divide directly with surface lowering. If this were true one might expect to see an expression of ridge B/C all the way to the Whitmore mountains. The present topography (figure 2) shows that this is not the case; the ridge extends from the summit to less than one half the way to the divide. The effect of this problem is to perhaps severely underestimate the duration of flow from the Whitmore mountains.

The convoluted contours and relatively shallow depth of the saddle on ridge B/C suggest that the flow off the summit of ridge B/C may be a relatively recent feature related to the emplacement of the ice streams. To test this, a new model was set up to reflect this configuration. The bore hole site was moved to the summit of the ridge and a permanent flow line was mapped back to the Whitmore Mountains ignoring the saddle all together. The spreading parameter changed through time from very non-divergent ($R=10^7$ km) all over to very non-divergent only at the divide and linearly becoming divergent as the bore hole ($R = 25$ km) is approached. The ice sheet was permitted to thin according to the Denton and Hughes, and Robin and Drewry models while flow was sustained from the Whitmore mountains.

The net effect of all this is to over-estimate the duration of flow from the Whitmore Mountains by the length of time equivalent to actual the emplacement of the dome on ridge B/C. Which in the first two dynamic models was taken, rather arbitrarily, to be -4,000 a. However, if the emplacement of the dome was in fact a more recent feature then sustained flow from the Whitmore mountains is a more realistic view. Additionally, if flow-line capture were the mechanism for such change then this change may have occurred in a virtual geologic instant rather than a slow linear movement from the Whitmore mountains to it's present position.

Thus we have two endpoint models of the thickness of the past ice sheet and two endpoint variations on the mechanics of the ice divide movement.

The plots of depth versus elevation (figure 12) and age versus depth (figure 11) are again easily distinguished from one and other. Equally important is that the these are also distinguishable from those of the previous model (compare figures

9,10 to 11,12); suggesting that the timing of the emplacement of the small dome may be discernible from elevation data. Again, the -18,000 a level occurs relatively high in the core; in both cases near 650 m.

SENSITIVITY OF THE RESULTS TO INPUT PARAMETERS

The sensitivity study is conducted by varying each parameter of interest by a physically significant margin and observing the results. The study was conducted around the Robin collapse using the model that maintains flow from the Whitmore Mountains. The parameters were changed and the stratigraphy was then examined at two selected core depths. Depths of 650 m and 350 m meters were chosen because, in the standard model, their age corresponds roughly to the timing of both the beginning and ending of the specified changes.

Table 1 shows a representative portion of the results of this study. In general, none of the parameters have a pronounced impact upon the elevation results; the greatest magnitude change is less then 10 percent. The results of the implied ages, however, are significantly more sensitive.

In general, the age at 350 m is much more strongly related to changes in the flow parameters than is the age at 650 m. Parameters that produce negligible changes in age at depth produce changes on the order of 30 percent at 350 m. However, the age at 650 m does appear to be strongly related to accumulation rate at the divide.

DISCUSSION AND CONCLUSIONS

In summary the outcome is quite encouraging. The results indicate the -18,000 a level occurs relatively high in the projected core (between -630 m and -750 m, most probably near 675 m). This important because, confidence in measured core properties is higher at shallower depths. This age (-18,000 a) is the age predicted for the maximum thickness of the ice sheet both by Robin and Drewry, and by Denton and Hughes. Thus there is a good potential for discriminating which reconstructed ice sheet is correct.

Of equal importance, an appropriately designed coring program could describe not only the magnitude of the collapse but also how and when the ice divide has shifted. Once developed, such a description how the ice sheet changed may permit the discussion cause for the change.

Finally, the small ice dome on ridge B/C was shown to likely be a local ice divide capable of sustaining flow.

REFERENCES

Bolzan, J., Whillans, I. M., Ice sheet changes in Northwest Greenland from the Camp Century ice core data. (In prep)

Budd W. F., Young N, W., (1983) Applications of modeling techniques to measured profiles of temperatures and isotopes. (In Robin. de Q., editor The Climate Record in Polar Ice Sheets 1983. Cambridge University Press. Cambridge, England. p. 150-177)

Dansgaard, W., Johnsen, S. J. (1969) A flow model and time scale for the core from Camp Century, Greenland Journal of Glaciology, 8, 215-23.

Denton G. H., Hughes T. J. Fastook, J. T., Stuiver M, 1981. History of the marine ice sheet in West Antarctica during the last Glaciation: A working hypothesis. (In Denton, G, Hughes T., editors, 1981. The last great ice sheets. John Wiley & sons. New York, New York. p. 319-436)

Drewry, D. J. (1979) Late Wisconsin reconstruction for the Ross Sea region, Antarctica. Journal of Glaciology vol. 24, no. 90, 231-244.

Drewry, D. J., Robin, G. de Q. (1983) Form and flow of the Antarctic ice sheet during the last million years (In Robin. de Q., editor The Climate Record in Polar Ice Sheets 1983. Cambridge University Press. Cambridge, England. p. 28-38)

Houghton, J. T. (1977) The physics of atmospheres. Cambridge University Press. Cambridge, England. 202 p.

Hughes, T. J., (1973) Is the West Antarctic ice sheet disintegrating? Journal of Geophical Research vol. 78, no. 33 pp.7784-7910

Hughes, T. J., (1977) West Antarctic ice streams. Reviews of Geophysics and Space Physics, vol. 15, no. 1,pp 1-45.

Jouzel J., et al.(1987) Vostok ice core: A continuous isotope temperature record over the last climatic cycle (160,000 years). Nature, vol. 329 pp 403-407.

Marsden, J. E., and Tromba A. J.; (1981) Vector calculus W. H. Freeman and Co., New York, New York. 591 p.

Muszynski, I., Birchfield, G. E., (1987) A coupled marine ice stream-ice dhelf model. Journal of Glaciology, Vol. 33, No. 113, pp 3-11

Paterson, W. S. B., (1969) The Physics of Glaciers. Oxford. Pergamon Press, 250 p.

Rose, K. E., (1982) Radio-echo studies of bedrock in southern Marie Byrd Land, West Antarctica. Antarctic Geoscience.

Rose, K. E., (1979) Charastics of ice flow in the Marie Byrd Land, Antarctica. Journal of Glaciology. vol 24, no 90, pp. 63-75

Robin, G. de Q., (1985) Contrasts in Vostok core - changes in climate or ice volume? Nature. vol 316, 15 August, pp. 578-579.

Robin, G. de Q. (1983) Ice sheets: Isotopes and temperatures (In Robin. de Q., editor The Climate Record in Polar Ice Sheets 1983. Cambridge University Press. Cambridge, England. p. 1-19)

Shabtaie, S., Whillans, I. M., Bently, C. R. Surface elevations on ice streams A, B, and C. (In Prep)

Thomas, R. H. (1976) Thickening of the Ross ice shelf and equilibrium state of the West Antarctic ice sheet Nature Vol 259, pp. 180-183.

Weertman, J. (1974) Stability of the Junction of an Ice Sheet and Ice Shelf. Journal of Glaciology. vol. 13, no. 65, pp. 3-11

Whillans, I. M., Bolzan, J., Shabtaie, S., (1987) Velocity of ice streams B and C, Antarctica. Journal of Geophysical Research. vol. 92, no. B9, pp 8895-8902.

Whillans, I. M. (1979) Ice flow along the Byrd Station strain network Journal of Glaciology. vol 24, no 90 pp. 15-28

Whillans, I. M. (1977), The equation of continuity and its application to the ice sheet near "Byrd" Station, Antarctica. Journal of Glociology. vol 18, no 80, pp. 358-371.

Whillans I. M. (1983) Ice movement. (In Robin. de Q., editor The Climate Record in Polar Ice Sheets 1983. Cambridge University Press. Cambridge, England. p. 70-77)

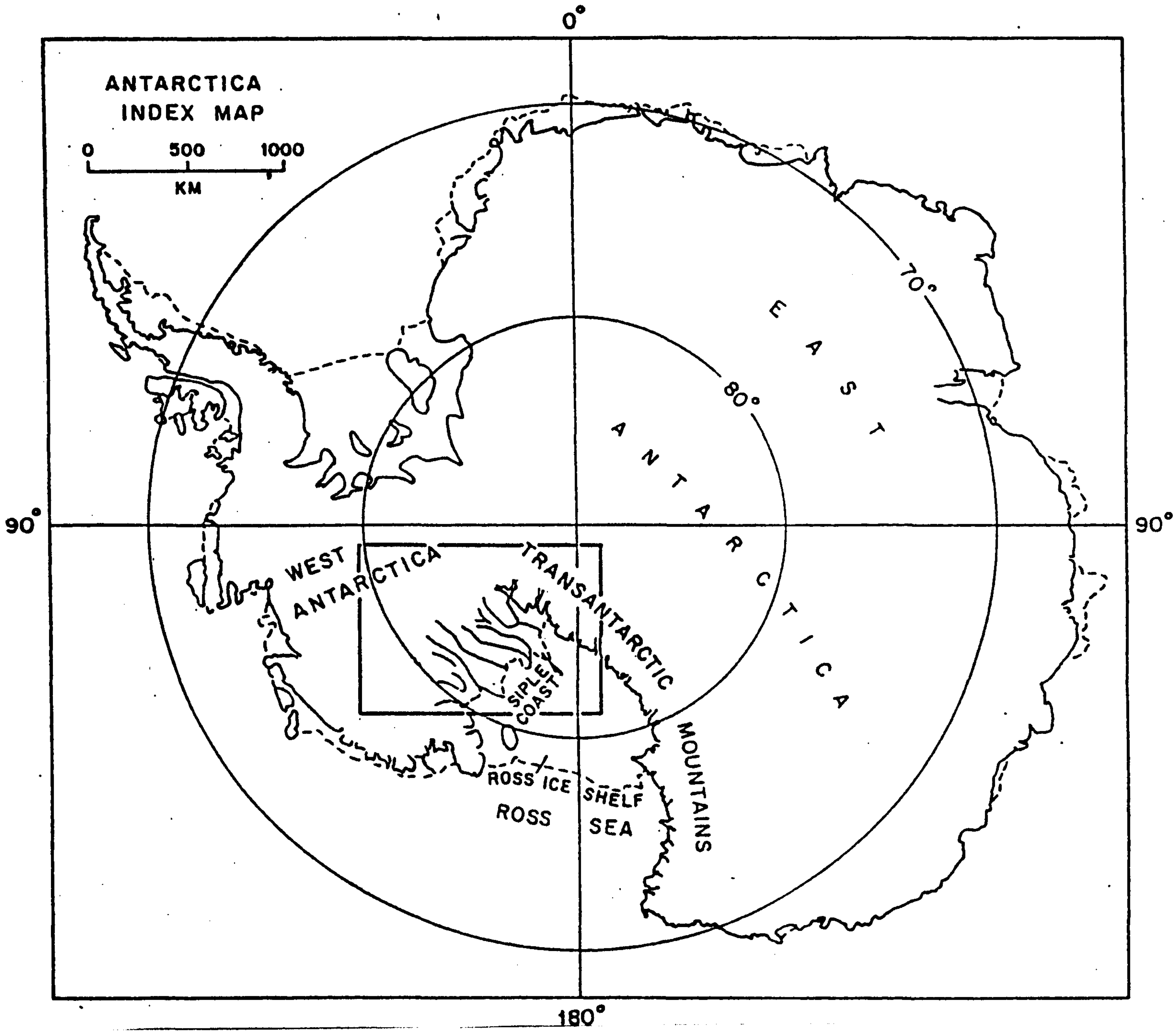


figure 1   Map of Antarctica showing Siple Coast from Shabtaie (unpublished)

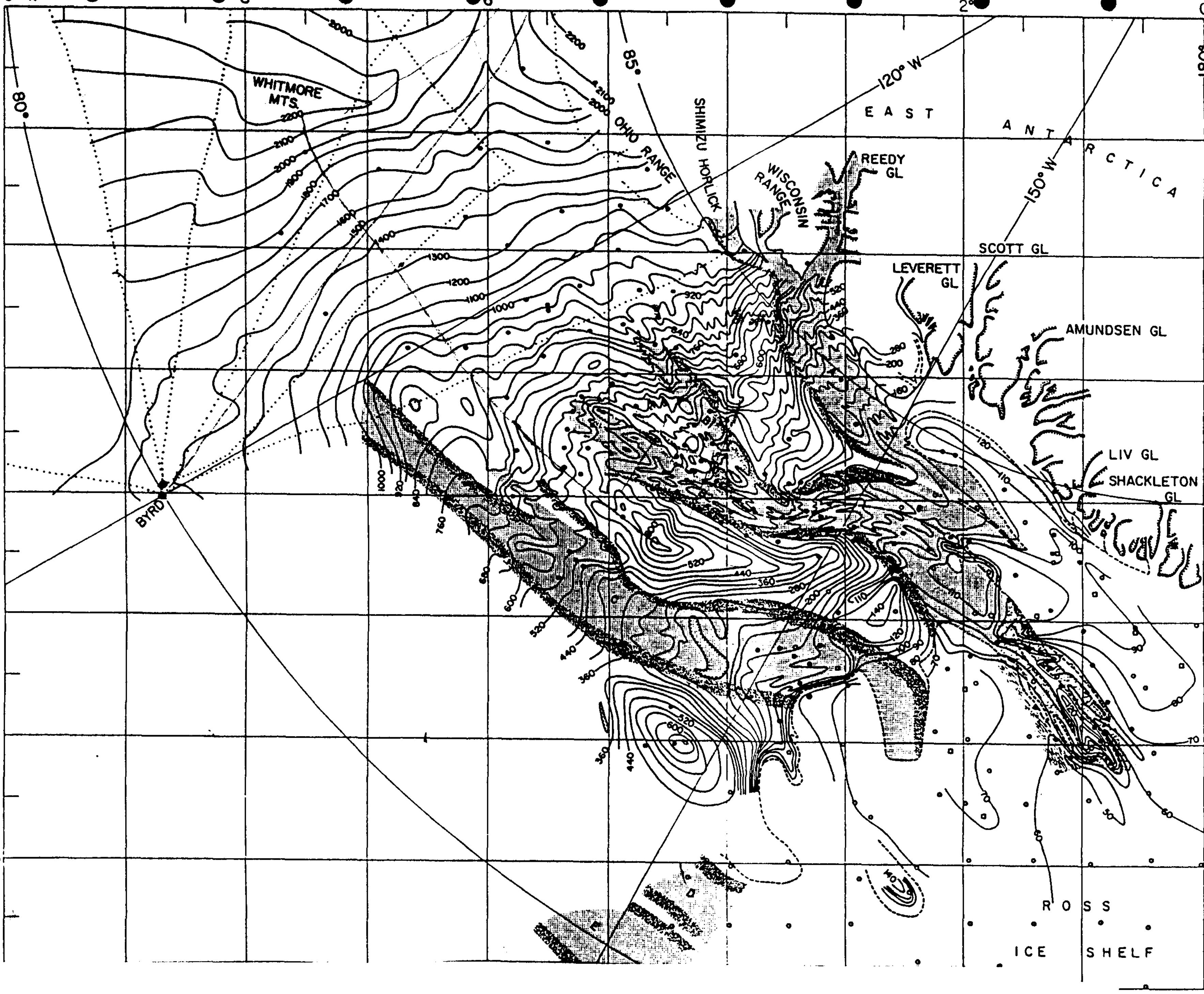


figure 2 Map of West Antartica showing ridge A/B, ridge B/C, ice streams A, B, and C, and The Whitmore mountains. From Shabtaie (unpublished).

Figure 2.16. Flowline profiles: Byrd Land.
Curves on right side.
(1) Present day.
(2) 18 000 a BP from Drewry's (1979) reconstruction and today's slopes.
(3) 18 000 a BP for expansion to edge of continental shelf and today's slopes.
(4) 18 000 a BP profile from Hughes *et al.* (1981).
Curves on left side.
(1) Present day.
(2) and (3) Suggested profiles at 18 000 a BP.

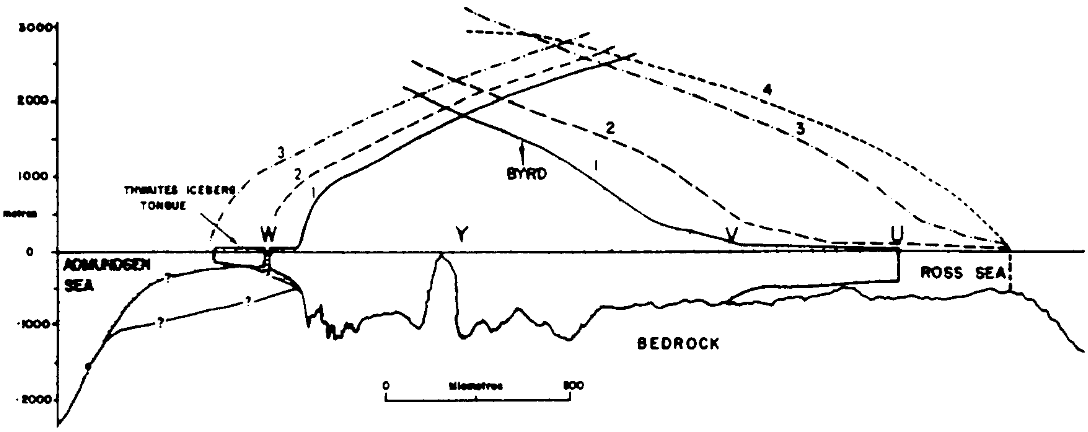


figure 3 Calculated surface profiles for both Denton & Hughes and Robin & Drewry reconstructions. From Drewry and Robin 1983.

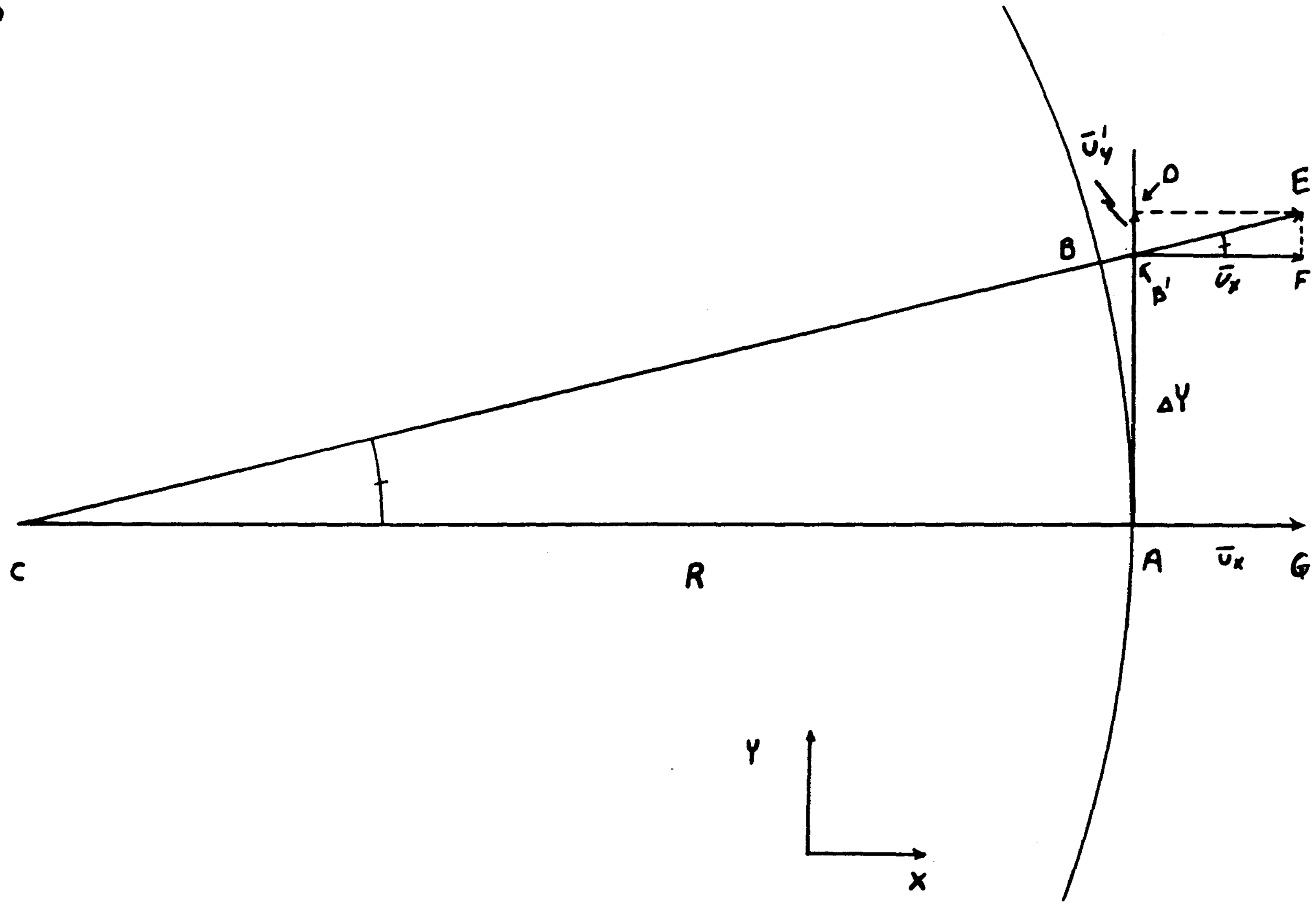


figure 4 Let $\overline{AG}$ be the average velocity along the flowline of interest $\bar{u}_x$. $\overline{BE}$ is a neighboring velocity vector call it $\bar{u}'$. However, for values of ΔY small compared R, $\overline{BE}$ may be approximated by $\overline{B'E}$. $\overline{B'D}$ would then correspond to the y-velocity component of u', or u'_Y , and $\overline{B'F}$ would be the x-componet u'_X . In the limit as Y tends to zero, and $\overline{u'_X}$ approaches $\overline{u}_X$. Also from geometry trangles CAB' and EDB' are similar. Therefore,

$$\frac{\Delta Y}{R} = \frac{\bar{u}_Y}{\bar{u}_x} \qquad \text{or,} \qquad \bar{u}_Y = \frac{\Delta Y \bar{u}_x}{R}$$

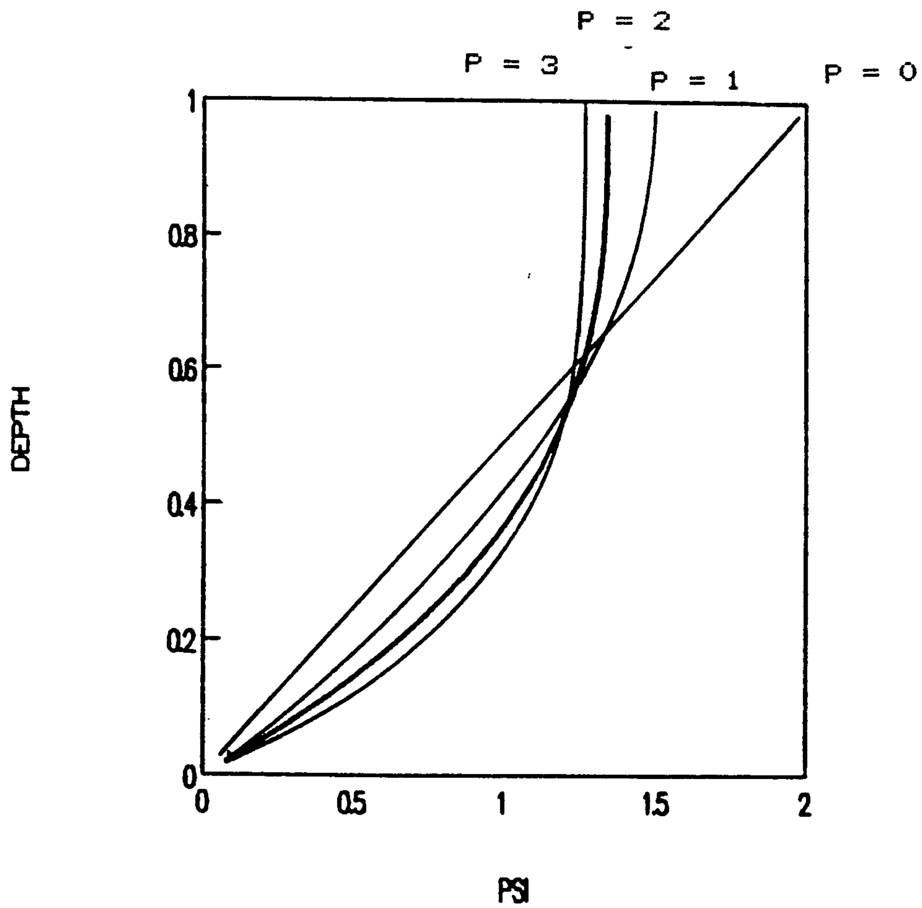


figure 5 The function PSI solved for normalized depth under various values for the parameter 'P'. 'P' values between 1 and 3 generally provide best fit for most measured profiles. A value of P = 2 was assumed throughout this study.

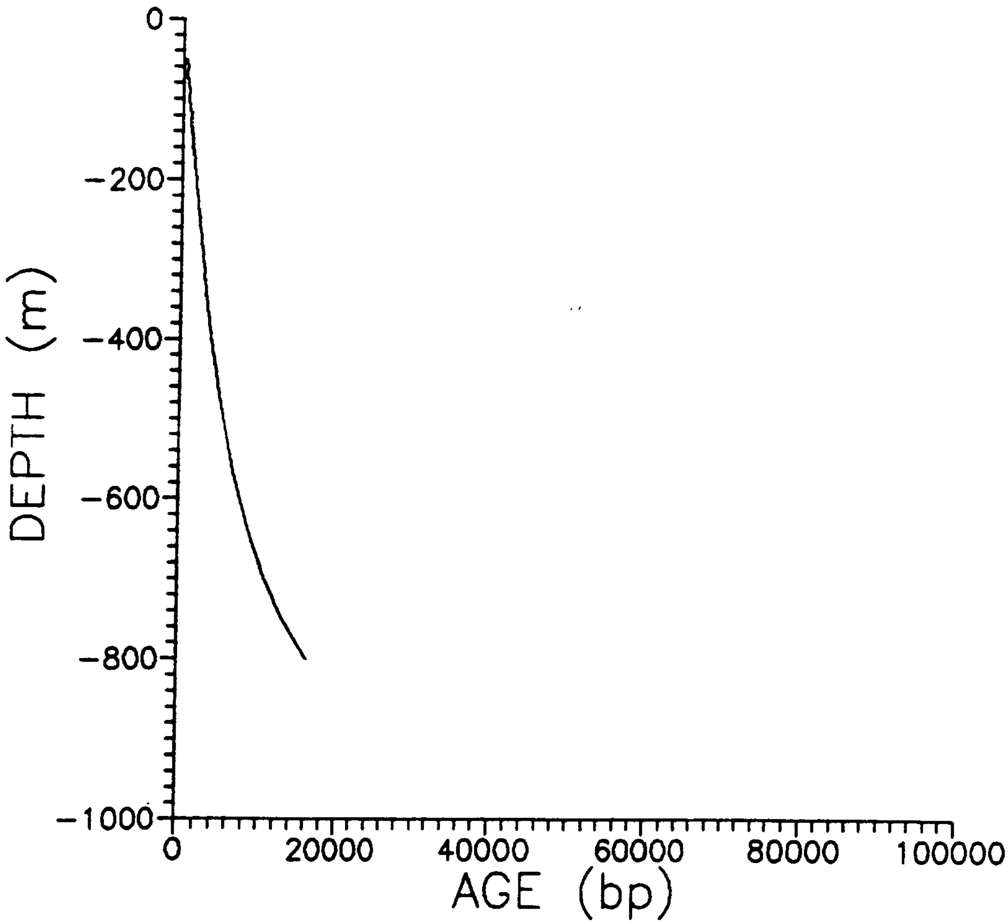


Figure 6 Plot of depth versus age from the steady-state scenario.

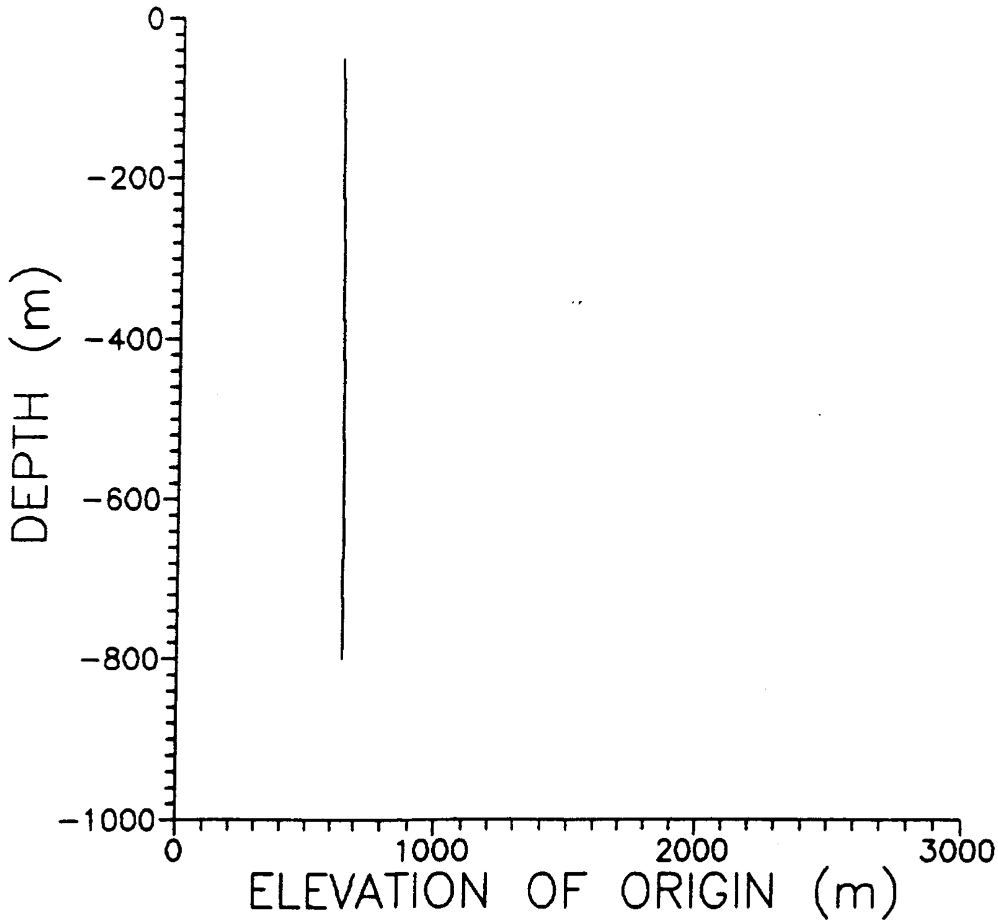
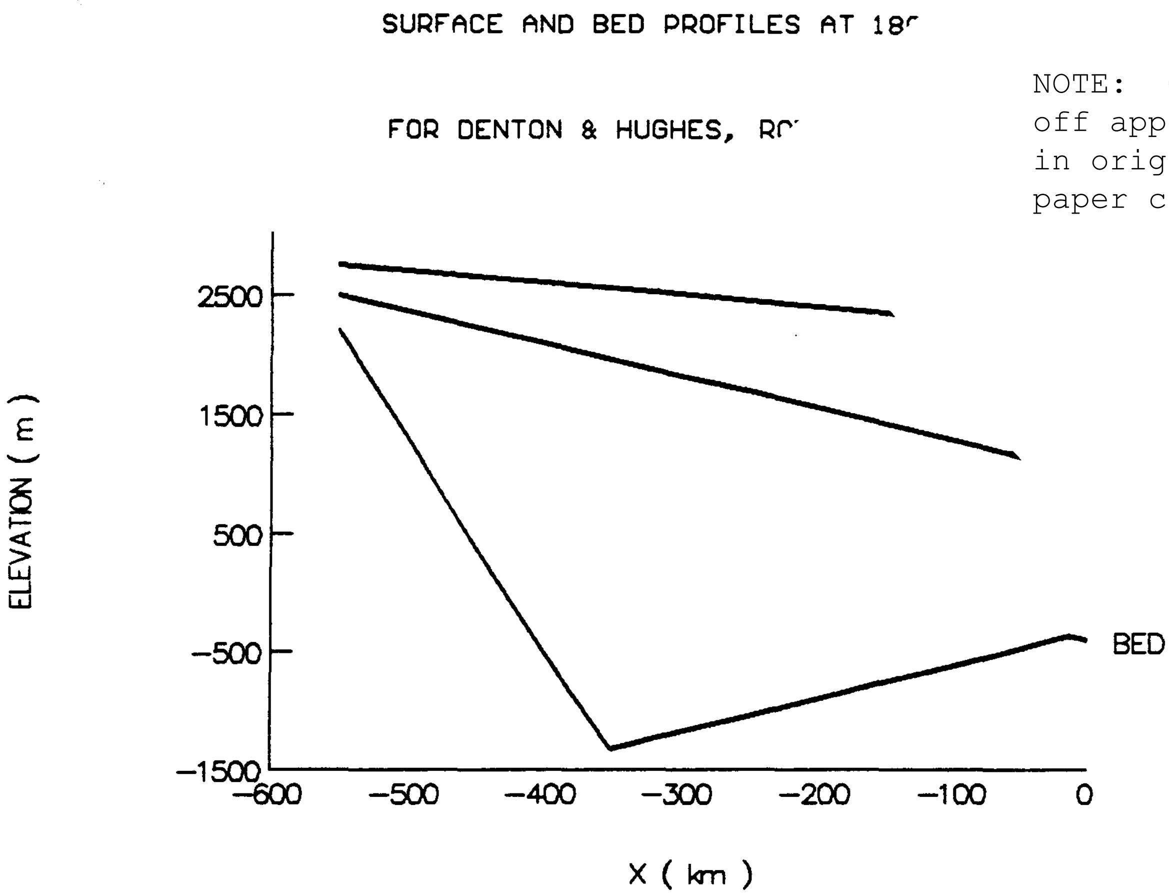


Figure 7 Graph of depth versus initial elevation for steady-state scenario.



NOTE: Cut off appears in original paper copy.

Figure 8 Surface profiles for times older than -18,000 years for Robin & Drewry and Denton & Hughes. Bed profile is not time dependent.

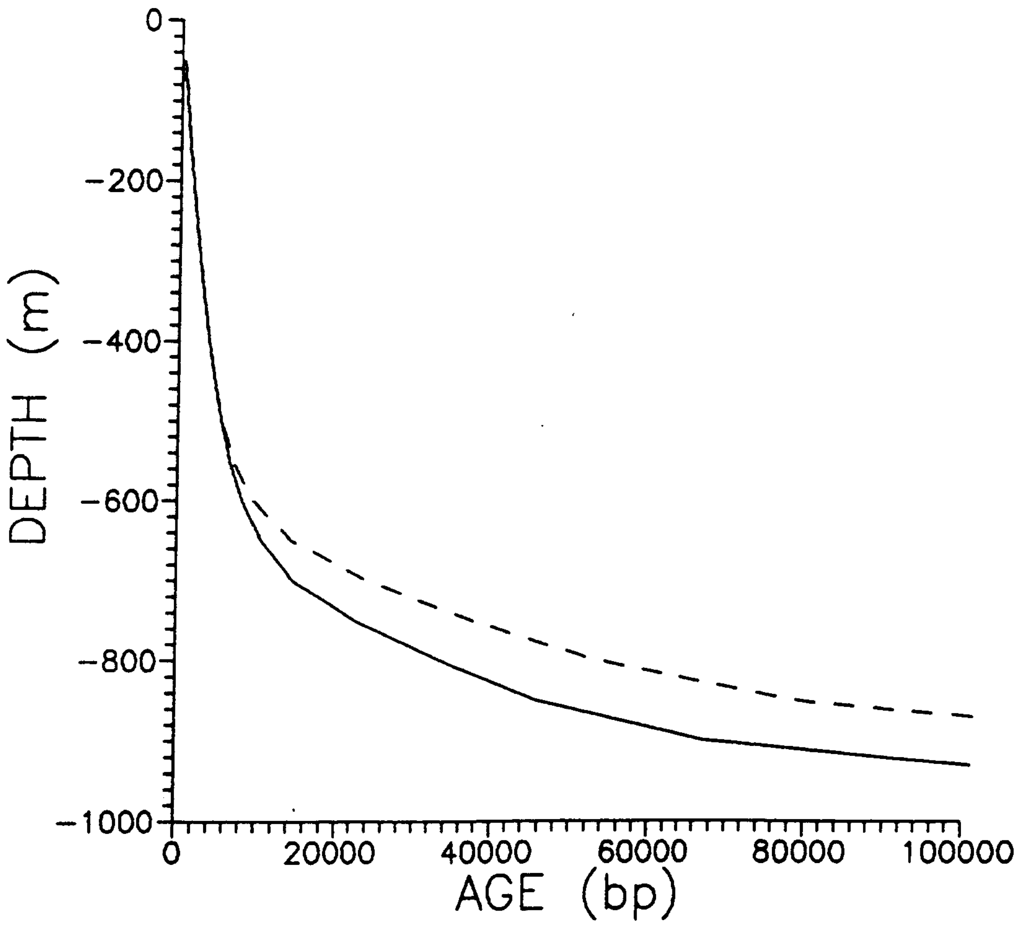


Figure 9 Graph of depth versus age for the linear movement of ice divide from Whitmore mts. to summit or ridge B/C from -18,000 to -4000 a. Dashed line is Denton & Hughes for ice sheet, solid for Robin & Drewry.

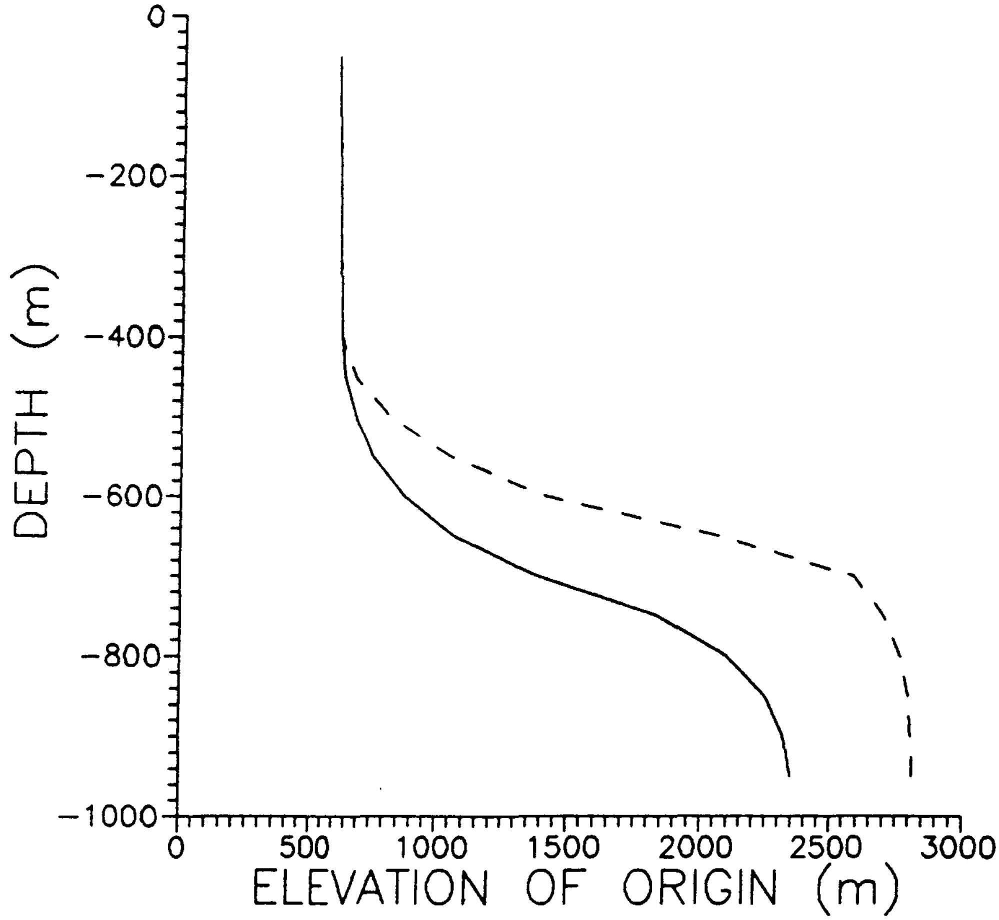


Figure 10 Graph of depth versus initial elevation for linear movement of the divide from Whitmore mts. to ridge B/C from -18,000 to -4000 a. Dashed line is for Denton & Hughes ice sheet solid for Robin & Drewry.

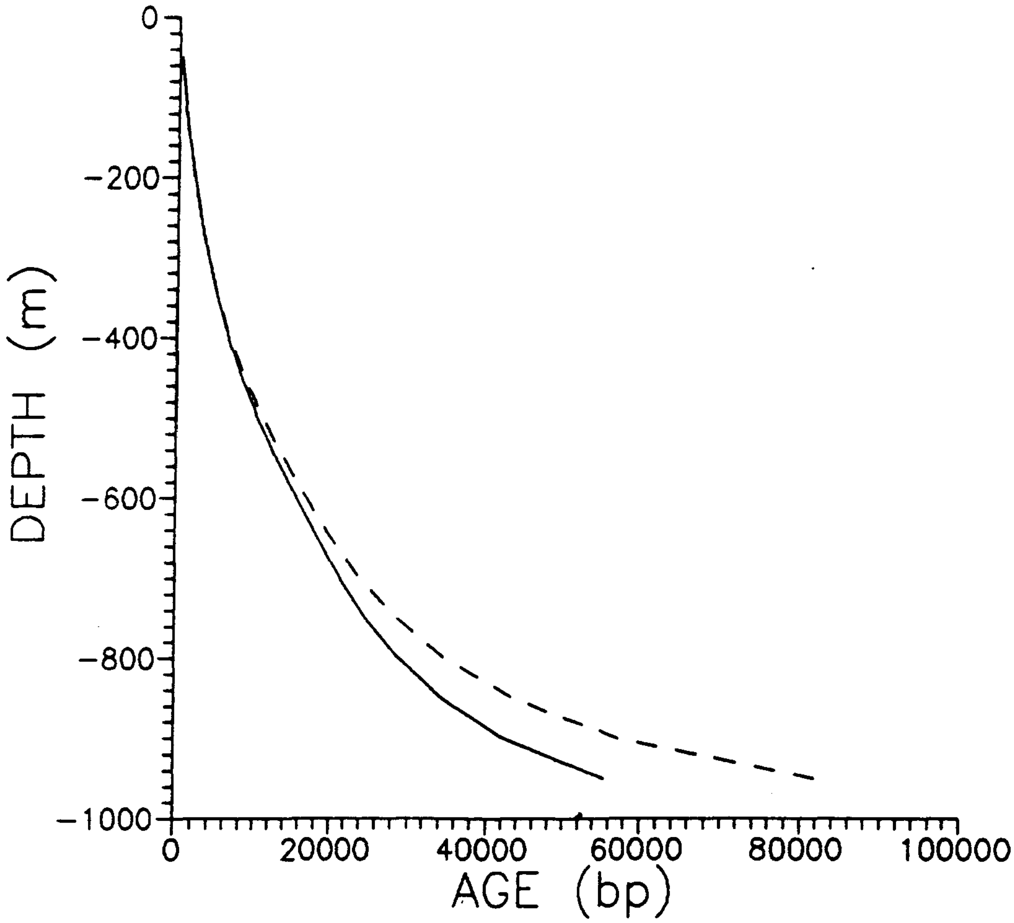


Figure 11 Graph of depth versus age for model of divide remaining at Whitmore mts. Dashed line for Denton & Hughes ice sheet, solid for Robin & Drewry.

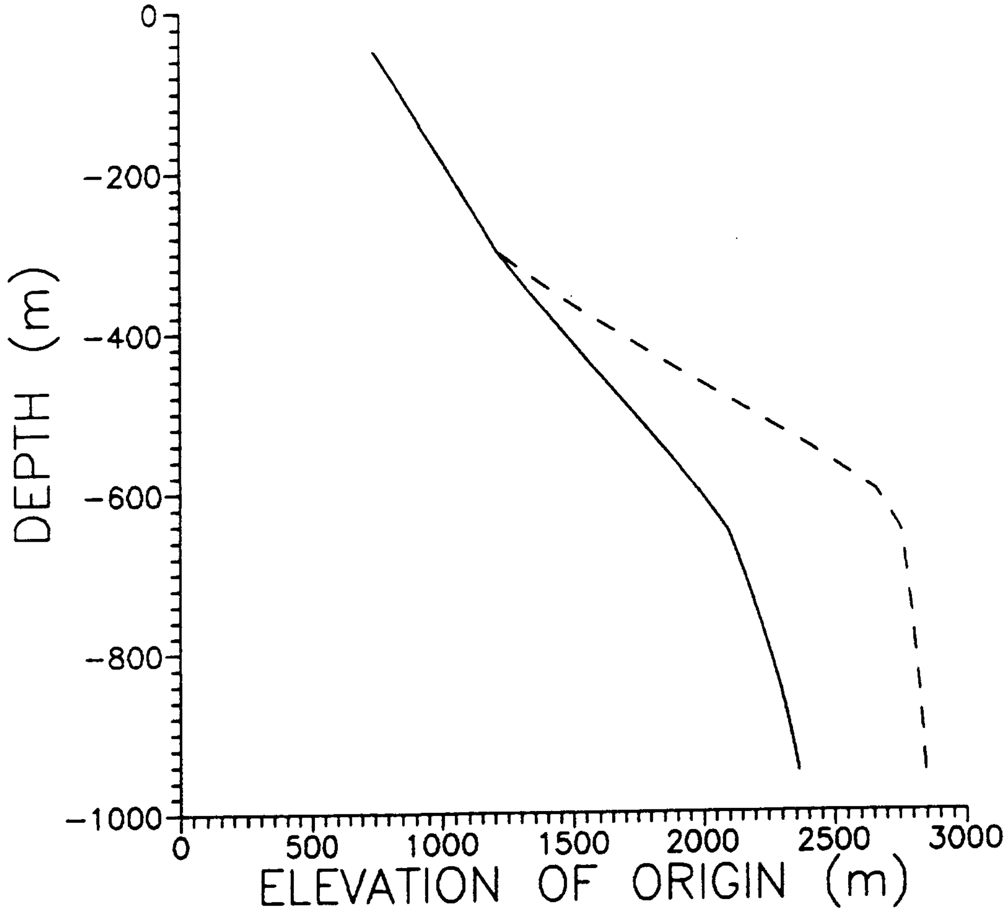


Figure 12 Graph of depth versus initial elevation for model of divide remaining at Whitmore mts. Dashed line is for Denton & Hughes ice sheet, solid for Robin & Drewry.

TABLE 1

SENSITIVITY STUDY

Results for standard model

DEPTH	AGE	INITIAL ELEVATION
650	-18263	2099
350	-3805	1211

PARAMETER CHANGED	STANDARD VALUE	NEW VALUE	NEW AGE	% CHANGE IN AGE	NEW ELEVATION	% CHANGE ELEVATION
P	2	1	-19773 -5387	-8.3 -41.6	2147 1408	-2.3 -9.4
P	2	3	-17407 -4754	4.7 -24.9	2051 1290	2.3 -3.8
$\dot{b}$ at bore hole (mm)	150	130	-19637 -5663	-7.5 -48.8	2115 1381	-0.8 -8.1
$\dot{b}$ at divide (mm/yr)	150	130	-19649 -5040	-7.6 -32.5	2083 1310	0.8 -4.7
$\dot{b}$ at divide (mm/yr)	150	190	-15952 -4815	12.7 -26.5	2225 1372	-6.0 -7.7
collapse timeing (yr)	-18000	-20000	-18184 -4980	0.4 -30.9	2067 1330	1.5 -5.7
collapse timeing (yr)	-18000	-16000	-18365 -4984	-0.6 -31.0	2099 1339	0.0 -6.1
surface slope 18 kbp	0.00245	0.0027	-18424 -4938	-0.9 -29.8	2223 1341	-5.9 -6.2
R at bore hole (km)	25	100	-18262 -3804	0.0 0.0	2099 1211	0.0 0.0
percent basal sliding	0	20	-17114 -4732	6.3 -24.4	2029 1286	3.3 -3.6